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The Role of The Heat Source of the Tibetan Plateau in the General Circulation[1]

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With 13 Figures

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Summary

In this paper, the thermal features of the atmosphere over the Tibetan Plateau in summer and their effects on the general circulation are reviewed. Some recent research results are reported. It is shown that the Plateau acts as a heat source in summer. Particularly the strong surface heating makes the air stratification very unstable and produces strong near-surface convergence and positive vorticity and upper layer divergence and negative vorticity. Intense convective activity generated thereby not only maintains such particular large-scale circulation pattern over the Plateau, but also transports large amounts of sensible heat, moisture, chemical pollutants, as well as air with low ozone concentration from near-surface layers to upper layers. A minimum centre of total ozone concentration and a huge upper layer anticyclone with a warm and moist core are thus observed over the Plateau in summer. The strong divergent flow and anticyclonic vorticity source in the upper atmosphere have a strong influence on the general circulation over the world via meridional as well as longitudinal circulations, and energy dispersion on a spherical surface. It is shown that the surface sensible heating of the Plateau is essential for the occurrence of the abrupt seasonal change of the general circulation there, and for the persistent maintenance of the Asian monsoon. It is also reported that the elevated heating of the Tibetan Plateau together with its mechanical forcing cause the early onset of the Asian monsoon to happen over the eastern coast of the Bay of Bengal, which then generates a favorable circulation background for the monsoon onset over the South China Sea. The Indian monsoon onset follows afterwards.

1. Introduction

The Tibetan Plateau covers nearly 2 and half million km^2, and its height reaches up to the middle of the troposphere. Such a huge block will evidently affect atmospheric motions significantly. Because in winter the Plateau is situated in the latitudes of the westerly jet stream and in summer it stands at the juncture of the westerlies and the easterlies, its role in the

[1] While I visited USA in the summer of this year (97) the sad news of the death of Professor Riehl came to me. This was a great shock to me. Herb was my esteemed colleague and friend. During the later half of my stay at the University of Chicago in the 40s I spent most of the time with Herb and worked with him. Every weak we have several discussion through which I learnt a lot from him. I cannot forget our discussions in one morning. This discussion helped me to formulate a paper "The circulation of the high troposphere over China in the winter of 1945–1946" which was published in Tellus (1950). This paper demonstrated for the first time the existence of a strong jet stream around the southern periphery of Tibetan Plateau. This jet stream is usually called southern jet stream in China, because there is also a northern one to the north of the Tibetan Plateau, These two jet streams merge into one downstream of the plateau forming the strongest jet stream in the northern hemisphere. This Tellus paper and a paper by Bolin, which also appeared in Tellus (1950), stimulated my interest in studying the role of the Tibetan Plateau in the general circulation for several decades. Because of Herb's stimulation, a colleague of mine and I write this review article in the volume in memorizing Herb's big contribution to meteorology.

[2] Because of the adoption of Pinyin in the 1960s, Yeh T. C. became Ye D.-Z

general circulation will have big seasonal variations as discussed by Yeh (1950).

Since the end of the 1940s, a series of studies have begun, aiming at understanding the effects of such a large highland on the general circulation, and many important results have been obtained. In particular, the dynamic effects of the Plateau in the winter half-year have been comprehensively examined (e.g., Charney and Eliassen, 1949; Bolin, 1950; Yeh, 1950). It has been found that the excitation of the quasi-stationary ultra-long waves in winter, the maintenance of the East Asian trough, and the formation of the Aleutian high in the stratosphere are all related to the existence of the Plateau. In the field of synoptic and dynamic analysis of plateau meteorology, many advanced studies have also been conducted in China. The book *Meteorology of the Tibetan Plateau*, written by Yang et al. (1960), was a survey of the research of the 1950's in China. In the period prior to the middle 50's, studies on the plateau meteorology were mainly concerned about its dynamical influence. However, in middle 50's Ye and his group (Koo and Ye, 1955; Yeh et al., 1957) and Flohn (1957) independently found that the Tibetan Plateau was a huge heat source in summer. Ye and his group grasped this important finding and then over a long period of time made a series of studies on the influence of this heat source on the general circulation (Koo and Yeh, 1955; Ye and Chang, 1974; Ye et al., 1974; Ye et al., 1979; Ye, 1982; Ye, 1993; Wu and Zhang, 1997; etc.). Others (Reiter, 1982; Flohn, 1968) also studied this topic. Since then besides the dynamical effects, the thermal influence of the plateau has also become a subject of intense study. Attention of these studies on the dynamic effect as well as thermal influence are not only confined to observations, but also include numerical experiments using GCMs as well as climate models. These studies aim to understand the mechanisms of the dynamical and thermal forcing separately, and also the linking of the two forcing effects of the plateau to the climate and its variability over the plateau, in the surrounding areas and even over the world. Besides, other meteorological phenomena over the Plateau were also investigated. In the 1970's, cooperative research was organized several times in China. Results suggested that the Tibetan Plateau area not only exhibits particular features of circulation, synoptic meteorology, and climate, but also that the Plateau plays different roles in the global general circulation during different seasons. A systematic summary of the studies conducted from the 1960's to the 1970's was published in the book *The Meteorology of the Qinghai-Xizang (Tibet) Plateau* by Ye, Gao, and colleagues in 1979.

In 1979, the First GARP Global Experiment (FGGE) was launched in which extensive data, including those from buoys, ships, aircraft, and satellites were collected. There was, however, not enough data concerning the Plateau. To fill this gap, the Qinghai-Xizang Plateau Meteorological Experiment (QXPMEX) program was conducted between May and August of 1979 in China. In this period, besides many new surface and pilot balloon stations, 6 radiative stations that had not existed before, and 4 new radiosonde stations in the central and western parts of the Plateau, were established. A large amount of data concerning radiation, heat balance, rainfall, clouds, and atmospheric parameters were obtained from these observations. With the efforts of meteorologists, the material has been subjected to careful analysis using advanced techniques, and many new results have been obtained. This greatly increased our knowledge of the Plateau's meteorology, especially in the summer season. The observational results and the associated analyses are summarized in the book *Advances in the Qinghai-Xizang Plateau Meteorology-The Qinghai-Xizang Plateau Meteorological Experiment (1979) and Research* by Zhang and his colleagues (1988). Systematic research on the thermal features of the Plateau and its effects on the general circulation of the global atmosphere based on these new observations were also made by Yang et al. (1990, 1992a, b, and c), Yanai et al. (1992) and many others (e.g., Reiter and Tang, 1984; Qian et al. 1984; Song et al., 1984).

In addition to observational studies, many numerical experiments by using general circulation models as well as climate models have also been conducted during the past years to understand the mechanism linking the mechanical as well as thermal forcing of the Plateau to the climate variability in the surrounding areas and over the world.

This paper presents a review of historical achievements in this field, with emphasis on more recent findings. In order to emphasize the thermal forcing of the plateau, attention is focused on the summer situation. The thermal features, the intense convective activity and its role in maintaining large-scale circulation patterns in the area are summarized respectively in Section 2–4. The impact on the global circulation of the strong upward motion and outflow in the upper layer of the Plateau resulting from the intense convective activities is presented in Section 5. Section 6 demonstrates how its elevated heating affects the abrupt seasonal changes of the circulation in the area, and the onset of the Asian summer monsoon and its persistent maintenance. The new findings of the existence of the minimum center of total ozone concentration over the Plateau is reported in Section 7. Finally, summary and discussion are given in Section 8.

2. Heat Source on the Tibetan Plateau

In the mid-1950's, Yet et al. (1957) and Flohn (1957) found independently that a huge heat source exists over the Tibetan Plateau in summer. At that time none of the two papers found this phenomenon through direct observations. However, in later years, due to the improvement in observational data, some quantitative analysis of the heat source over the Plateau appeared to be possible. Using climate data, Zuo (1963) estimated the effective surface radiation in Lhasa for the four seasons. Chen et al. (1965) calculated the heat source intensity over Lhasa by means of radiosoundings. Based on these an various other kinds of data the values of overall heat source intensity for different months over the eastern and western regions of the Plateau, separated by 90° E, were given by Ye et al. (1979) for a 10-year average. Table 1 gives the average intensity of the heat source over the plateau for each month of the year. In Table 1 E is the heat given to the atmosphere calculated by the following equation

$$E = SH + LR_1 + LP + SR - LR_2$$

where SH is the turbulent sensible heat transfer from the ground; LR_1, the effective radiation from the ground; LP, the latent heat released from precipitation; SR, the short wave radiation of sun light absorbed by the atmosphere; and LR_2, the outgoing long wave radiation at the top of the atmosphere.

However, the calculations of this table met challenges. From May to August of 1979, a large-scale Tibetan Plateau observation experiment (QXPMEX) was carried out. Among other things the radiation and heat balance of different underlying earth surfaces on the Plateau were observed, from which the distribution of the heat source during this period could be calculated. A similar observational experiment was carried out from August 1982 to July 1983 at four typical surface stations (Gerze, Nagqu, Lhasa, and Ganze) by the Lanzhou Institute of Plateau Atmospheric Physics. Based on these observation the heat source on the Plateau had been recalculated (Zhang et al., 1988). The recalculated heat source was much smaller, only about half of that by Ye. The discrepancy mainly comes from the term $SH = C_p \rho C_D V(T_s - T_a)$. The meanings of these symbols are those generally used in meteorology. The choice of the value for C_D is the trouble. In general the value of C_D increases with the roughness of the ground surface. Ye in his paper gave a brief discussion about how it varies from different authors' estimates. Here we just give some examples. For the terrain of medium high mountains in Siberia some author's (see Cressman, 1960) assumed C_D to be

Table 1. *Annual Variations of the mean Heat Source of the Atmosphere over the Tibetan Plateau (E), and the associated Surface Sensible Heat-Flux (SH), Surface Net Radiation Flux (LR_1, Lhasa only), and Latent Heat Release due to Condensation (LP). Unit is Wm^{-2}. Reproduced from Ye et al., (1979)*

Heat source term	Jan.	Feb.	Mar.	Apr.	May	Jun.	Jul.	Aug.	Sep.	Oct.	Nov.	Dec.	Ann. mean
SH (whole Plateau)	21	43	79	124	146	141	116	96	80	63	32	13	80
LR_1 (Lhasa)	81	81	81	81	81	54	54	54	79	79	79	81	74
LP (whole plateau)	4	6	19	16	25	63	78	78	44	16	3	1	29
E (whole Plateau)	−72	−42	25	60	94	109	101	74	44	−10	−54	−77	21

4.0×10^{-3}. Cressman (1960) gave $C_D<8\times10^{-3}$ for the Rocky Mountains and the Tibetan Plateau. Based on observations, Zeng and Guan (1975) gave $C_D=32\times10^{-3}$ for the terrain of broken rock near the foot of Mount Qomolangma. Fiedler and Panofsky (1972) estimated the roughness parameter Z_0 to be 0.99 m and 1.65 m for middle and high mountains respectively. If these values of Z_0 are used to calculate C_D, it would be very large. Considering the wide rang of values for C_D for different kind of terrain, Ye estimated C_D over the plateau to be $6\text{–}10\times10^{-3}$ and he used 8×10^{-3} for C_D to calculate the overall heat source intensity of the Tibetan Plateau. But C_D used by Zhang et al. (1988) was about $4\text{–}5\times10^{-3}$ based on the observations at several stations in Tibet. We do not know what the terrain of these stations looks like. One could imagine that the stations chosen for observations were situated at relatively flat places (although not flat in terms of large areas) for ease of observation. Thus, the use of their value of C_D to represent the overall value of C_D for the plateau could be strongly questioned.

Further it was also observed (Ye and Gao, 1979) that in summer the value of T_s-T_a increased very rapidly with height of the mountain. At the height of 4950 m of Rongbu Temple, its monthly mean in June reached values as large as 11 °C. At the height of 8 km of Mount Qomolangma, its value already exceeds 17 °C in May. Also considering the high wind speeds at the top of the mountain, one can easily imagine that high mountains may behave like a chimney shooting large amounts of heat to the upper troposphere. The number of mountains over the plateau is quite large. Their chimney effect will undoubtedly increase the intensity of the heat source of the plateau in summer. Considering all these factors the discrepancies between Ye's and the other authors' calculation of the heat source intensity may easily be explained.

Furthermore, using the sounding data of the added upper air stations of QXPMEX of 1979, Nitta (1983) and Lou and Yanai (1984) calculated the vertical distribution of such heating and moisture in the atmosphere over the plateau. In addition to it, the heat source intensity of the plateau was also computed. Their computed intensity agreed quite well with that by Ye. Using the ECMWF/TOGA complementary data set, Wu and Zhang (1997) calculated the longitude-time distribution of surface sensible heat flux meridionally averaged between 27.5 and 37.5° N in 1989. It was found that from May 1989 onwards over the Plateau area SH varies from 100 to 225 W m^{-2}. However the NCAR/NCEP monthly mean reanalysis data from 1980 to 1995 give the SH value less than 100 W m^{-2} over the Plateau area during the summer months (Liu Yiming, 1997, personal communication). These findings imply that the accurate estimation of SH over the Tibetan Plateau is still a complicated and unsolved problem. But the magnitude of SH provided by Zhang et al. (1988), which was obtained by using small values of C_D, is the least. We note that the methods of estimates by Ye, Nitta, and Lou and Yanai and of reanalysis by using numerical models were totally different. We may say that the heat source intensity given by Table 1 is believable to the order of magnitude.

According to Table 1 the average heat source of the atmosphere over the whole Plateau is positive during the summer months, with a maximum in June around 109 W m^{-2}, and negative in the winter months with a minimum in December around -77 W m^{-2}. The intensity of the heat source in summer over the western part of the plateau is much stronger than over its eastern part, and the atmosphere over the western part starts to be heated about one month earlier and cools off about one month later than the atmosphere over the eastern part.

3. Convective Activity over the Plateau in Summer

Strong solar radiation produces strong surface heating, which in turn gives rise to a large temperature lapse rate near the surface of the Plateau in summer. In Lhasa, from the surface to 470 hPa, the 10-year daily mean atmospheric lapse rate in July exhibited super-moist-adiabatic characteristics, or close to dry adiabatic characteristics (Ye et al., 1974). During the QXPMEX period in 1979, in the daytime, and especially in the afternoon, a super-dry-adiabatic lapse rate usually occurred in the layer 1–2 km from the ground, and could persist for up to 6 hours. Further, a super-dry-adiabatic lapse rate layer as thick as 2.3 km was observed, with the maximum

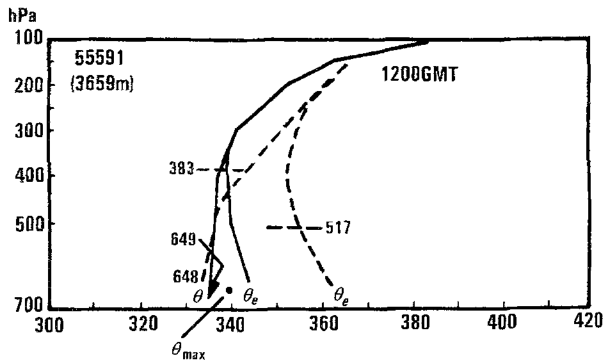

Fig. 1. Vertical profiles of 10-day mean potential temperature (θ) and equivalent potential temperature (θ_e) over Lhasa before (solid curves) and during (broken curves) the rainy season in 1979. Adopted from Luo and Yanai (1984)

lapse rate reaching −13.9 C/km in a layer 1.5 km in thickness. Thus, a strong lapse rate must be accompanied by intense convective activity, resulting in a deep mixed layer over the Plateau. Luo and Yanai (1984) studied the thermodynamic structure of the atmosphere over the Plateau before and during the rainy season of 1979. Figure 1 shows the 10-day mean vertical distribution of potential temperature (θ) and equivalent potential temperature (θ_e) before and during the rainy season of 1979 at Lhasa that they found. The mean surface pressure and the lifting condensation level at 1200 GMT are also given in this figure. It can be found that before the rainy season, the top of the mixed layer is at 400 hPa, with a potential temperature of 336 K. This means that air on the surface can be lifted to a height even higher than 400 hPa, carrying surface sensible heat into a layer deep in the troposphere. By comparing the θ and θ_e curves before and during the rainy season, it can be found that the atmosphere is more stable with dry convection but more unstable with moist convection within the rainy season. During the rainy season, due to the appearance of a deep unstable layer and the low condensation level (517 hPa) above the ground (mean surface pressure about 600 hPa), air from the surface could reach 200 hPa after condensation, carrying water vapor to the upper troposphere and increasing the air temperature there by condensation (Ye et al., 1974; Luo and Yanai, 1984).

The unstable stratification over the Plateau undoubtedly causes strong convective activity.

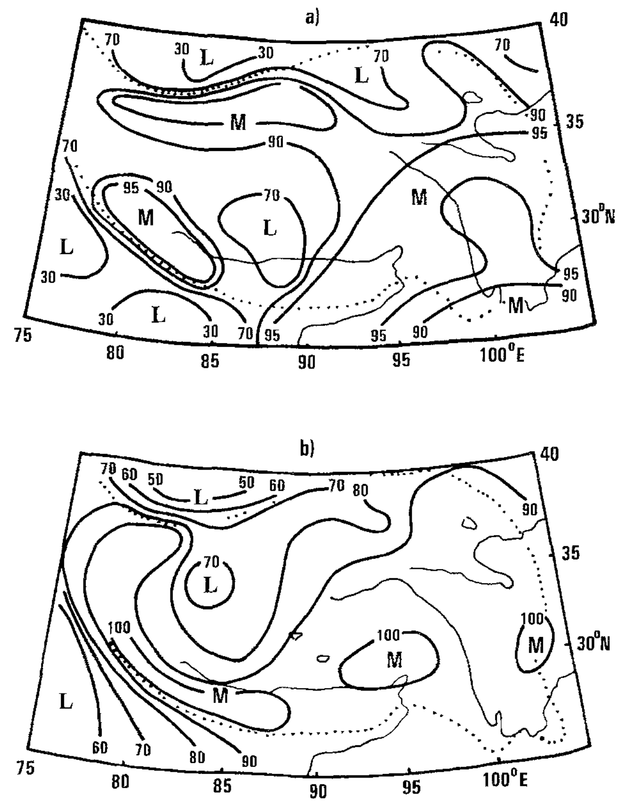

Fig. 2. Frequency distribution of cumulus averaged from 1300 to 1700 h (Beijing time) in 1979 before (a) and during (b) the rainy season. Adopted from Qian et al. (1984)

This has been noticed and has been the subject of discussion for a long time (Flohn, 1968; Ye et al., 1974; Yuan, 1979). Based on QXPMEX data obtained in 1979, Qian et al. (1984) gave a detailed discussion on such convective activity. Figure 2 shows the frequency distribution of convective clouds over the Plateau from 1300 to 1700 (local time) (a) before and (b) during the rainy season in 1979. In the afternoon, the distribution frequency is greater than 70% even before the rainy season (a) in most parts of the Plateau. The strongest convective activity took place over the southeastern Plateau, the western Himalayas, and Kunlun Mountains. The weakest convective activity was over the Shen Zha region and the Qaidam Basin. During the rainy season, convective activity increased to a frequency of 90% in most parts of the Plateau, and even the minimum frequency was not less than 70%. It was also found that among the convective clouds cumulonimbus was dominant. Using three years of data, Yuan (1979) counted the number of days in which cumulonimbus appeared at all stations of the Plateau for different months. He found that the minimum number, occurring in Galmud (32°30′ N, 80°05′ E) in July, was 16 days. At all other stations the number of days was greater than 24. These data, together with Fig. 2, illustrate that the Plateau in summer is among the regions of greatest connectivity in the world. In fact, even in winter, there is still considerable convective activity over the Plateau.

It was long believed that the convective activity should be much more pronounced over the eastern part than that over the western part. This impression may come from lack of data in the west in past years. The enriched data in 1979 showed that this may not be the case. What is noticeable from Fig. 2 is that convective activity over the western Plateau is not much less than that over the east. More detailed studies on this phenomenon were conducted by Yang et al. (1992b) based on the data set offered by Professor Zhu Fukang of the State Meteorological Administration who combined FGGE data with QXPMEX data to produce a new data set, greatly improved in quality, especially for the western Plateau. Yang et al. computed the distribution of vertical motion (Fig. 3) and compared it with that of convective clouds estimated from satellite cloud pictures (not shown), and found that in both the eastern and western regions of the Plateau, there is a center of maximum upward motion coinciding with the high frequency center of convection clouds. Using satellite cloud pictures taken in August of 1975, the Research Group on Experimental Simulation (1977) calculated the percentage of cloudless days and pointed out that the minimum center was around

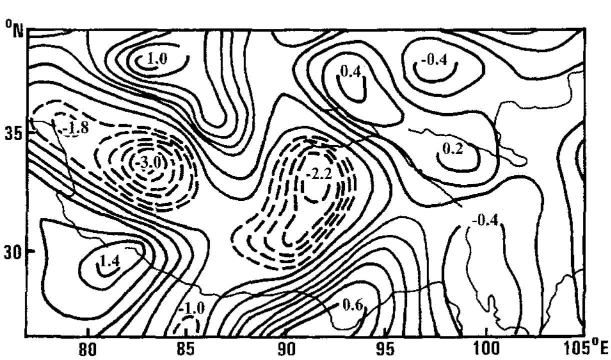

Fig. 3. Distribution of the vertically averaged "vertical velocity" ω at 1800 h (Lhasa local time) in July 1979. Unit is in 10^{-3} hPa s^{-1}. Adopted from Yang et al. (1992b)

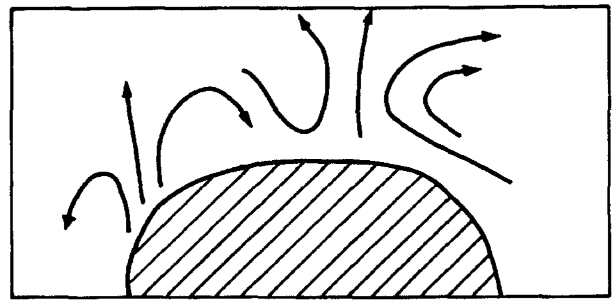

Fig. 4. Schematic diagram of vertical circulation observed at a heated "Plateau" (sketched region) in an annulus experiment. Adopted from Research Group at IAP (1978)

31° N, 79° E over the western Plateau. This center was distinctly separated from the eastern center. This can be considered as further evidence of the existence of an ascending center over the western Plateau.

All of the above facts show that the previous conclusion of latent heat being small in the western Plateau and that sensible heat is dominant (Flohn, 1968; Ye, 1979) should be corrected to some extent. The former conclusion was drawn from precipitation data. As there were only very few observation stations in the western part of the Plateau, intense rainfall might not have been observed.

It is interesting to note that the intensity of convection is not uniform over the plateau, but concentrated in two regions, one in the east and the other in the west. In a laboratory experiment, the Research Group on Experimental Simulation at IAP (1978) simulated the convective systems caused by heating of an oblong "plateau" in a rotating annulus. Surprisingly enough, although the oblong "plateau" was made quite uniform and smooth, two convective centers were found (Fig. 4) just over the "plateau", as observed in the real atmosphere. The horizontal scale of the convection was calculated and found to be consistent with the inference in the Chandresekhas' (1961) theory.

4. Interaction between Convective Activity and Large Scale Circulation

It is well-known that in summer over the Plateau and its surrounding area there is a large high pressure system in the upper troposphere and stratosphere, but a strong cyclonic circulation in the surface layer of the Plateau. What we are concerned about in this section is the relationship between the small-scale convective systems and the very large-scale circulation systems. Ye et al. (1974) pointed out that convective activity can bring water vapor to the upper troposphere where air is subjected to condensation heating, leading to the maintenance of high temperature and humidity there. Based on the characteristics discovered by Krishnamurti (1971) that the center of the streamline pattern almost coincides with that of the velocity potential at 200 hPa over the Plateau, Ye et al. (1974) examined the vorticity balance, and found that convective activity is an important factor in maintaining that balance in the whole atmosphere over the plateau (cyclonic vorticity near the surface produced by convergence, and anticyclonic vorticity in the upper troposphere and lower stratosphere produced by divergence). This result is similar to that obtained by Holton and Colton (1972). However, Murakami (1987) pointed out that the equiscalar lines of the stream function and velocity potential are not coincident, but intersect each other at large angles at 200 hpa in summer. In addition, Sardeshmukh and Held (1984), and Sardeshmukh and Hoskines (1985) suggested that nonlinear advection of vorticity can compensate for the effect produced by the divergence term, resulting in the balance of vorticity in the upper layer of the atmosphere at low latitudes.

In view of these differences, using the best available data set derived from FGGE and QXPMEX data from May to August 1979, Yang et al. (1992c) reexamined the problem of vorticity balance in the atmosphere over the plateau by computing every term of the vorticity equation. The mean vorticity equation can be written as:

$$\frac{\partial \bar{\zeta}}{\partial t} + \bar{V} \cdot \nabla \bar{\eta} + \bar{\omega} \frac{\partial \bar{\eta}}{\partial p} + \left(\frac{\partial \bar{\omega}}{\partial x} \frac{\partial \bar{v}}{\partial p} - \frac{\partial \bar{\omega}}{\partial y} \frac{\partial \bar{u}}{\partial p} \right) + \bar{\eta} \bar{D} + TT + R = 0$$

where

$$TT = \overline{v' \cdot \nabla \eta'} + \overline{\omega' \frac{\partial \eta'}{\partial p}} + \overline{\left(\frac{\partial \omega'}{\partial x} \frac{\partial v'}{\partial p} - \frac{\partial \omega'}{\partial y} \frac{\partial u'}{\partial p} \right)} + \overline{\zeta' D'}$$

Here "$-$" denotes monthly average; "$'$" is the deviation from the monthly mean; "TT" is the sum of all the transient terms; "R" is the residual term for maintaining the balance of the equation, which is mainly determined by subgrid-scale convective activity. The calculation were made for two domains with ground height respectively higher than 4500 meters and 2750 meters over the Plateau, with a mesh scale of 1.875 longitude $\times 1.875$ latitude. In order to study diurnal variations, data taken twice a day (at 0600 and 1800 local time) were employed. Area mean results were obtained both for the whole Plateau and for its eastern and western parts separated by 90° E.

Figure 5 shows results calculated for the region of the plateau above 4500 meters. The main terms are the divergence term ($\bar{\eta}\bar{D}$), the advection term ($\bar{V} \cdot \nabla \bar{\eta}$) and the residual term (R). All the other terms were also computed, but they were too small compared with the main term and are not drawn in the figure. For better comparison of the two terms, $\bar{\eta}\bar{D}$ and $\bar{V} \cdot \nabla \bar{\eta}$, it is seen in these figures that the curve for R is only show in Fig. 5c. The general circulation of the atmosphere still exhibits winter characteristics in May, the vorticity budget (Fig. 5a) shows the balance of nonlinear advection and the divergence term to be dominant in these two areas (>2750 and >4500 meters) with R over small. In June, the East Asian circulation undergoes an abrupt seasonal change and the summer circulation begins. The convection term (R) is then enhanced obviously. In the morning at 0600 (local time), it can already be seen that the divergence term cannot compensate the effects of the advection term alone. At 1800, seen from the right panel of Fig. 5b, the advection term becomes much smaller than the divergence term which must then be balanced mainly by R. The convection term (R) increases further at both 1800 and 0600 in July. In the morning (Fig. 5c, left) the convergence layer producing cyclonic vorticity was located under 480 hPa, with a thickness of approximately 100 hPa (the mean surface pressure of the Plateau in the area over 4500 meters is approximately 600–550 hPa). The divergence layer exists above 480 hPa, producing anticyclonic vorticity. The left panel of Fig. 5c shows that in the lower layer the subgrid convection term (R) is larger than the nonlinear

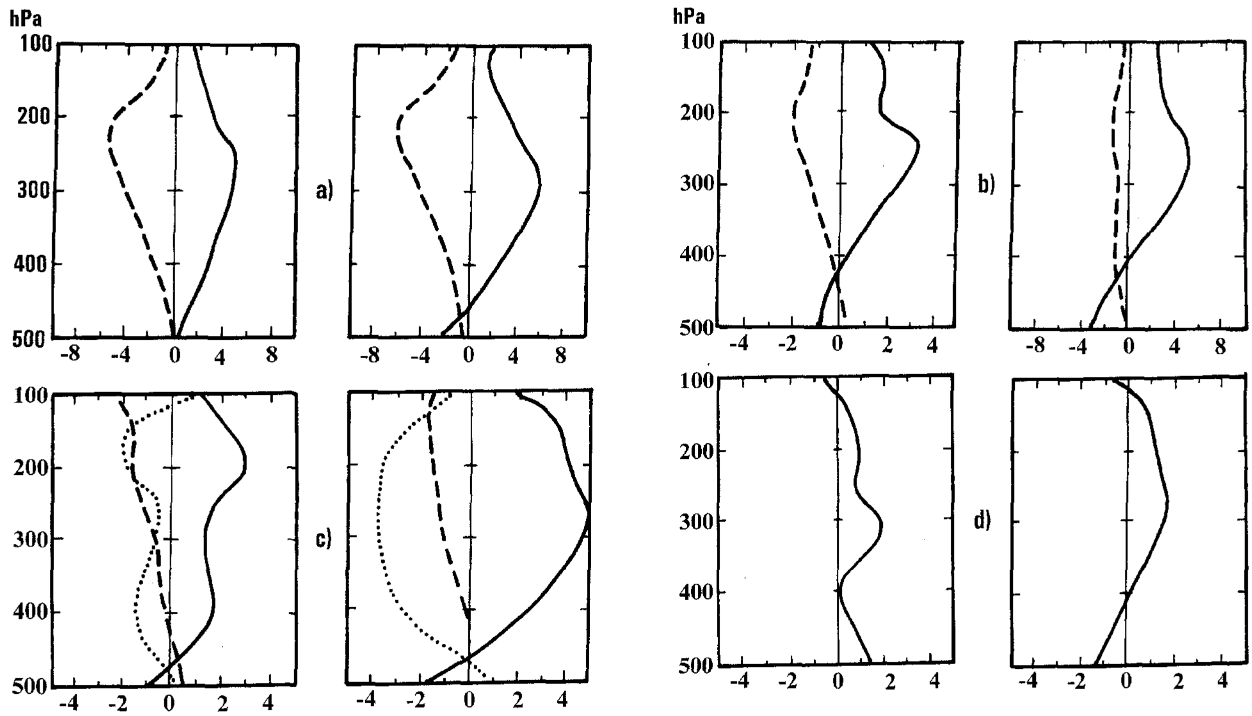

Fig. 5. Vertical distribution of divergence term (solid line), advection term (dashed line), and subgrid-scale term (dotted line) in the monthly mean vorticity budget in 1979 for the Tibetan area over 4500 m. Data obtained at 0600 (left) and 1800 (right) local time. Unit: $10^{-10}\,s^{-2}$. (a) May; (b) June; (c) July; (d) August. Adopted from Ye (1993)

advection term $(\bar{V} \cdot \nabla \bar{\eta})$, it plays a more important role under 300 hPa. Above 300 hPa, the R term and the advection effect are almost equivalent. In the whole, these two terms can basically offset vorticity production by divergence. From morning to afternoon, both the R and the divergence term increase greatly (Fig. 5c, right) due to the full development of convective activity, which causes the strengthening of convergence at low levels and of divergence at high levels. As a result, the subgrid-scale convective term (R) is enhanced, which alone can basically balance the divergence term with the advection term much less than R. In August (Fig. 5d), both in the morning and afternoon the advection term is negligible and not shown in the figure, thus the R term plays a more important role than the advection term in the vorticity balance.

The mechanism that explains how the effects of the divergence field can be balanced by convective activity in summer is evident. Cyclonic circulation at low levels, and anticyclonic circulation at high levels constitute the basic vertical structure of the pressure configuration over the Plateau in summer. The cyclonic vorticity produced by convergence at low levels can be transported to high levels by the ascending air of small-scale convective systems to offset the anticyclonic vorticity produced by divergence there. The descending air of these convective systems carries the anticyclonic vorticity from high levels to low levels, cancelling low-level cyclonic vorticity. But in May the pressure systems over the plateau still posses winter characteristics with no stable cyclonic circulation at low levels and no stable anticyclonic system at high levels over the Plateau, thus, vorticity transportation by convection is very weak.

As early as in the later 1950's Yang et al. (1960) pointed out that the intensity of diurnal variation in meteorological elements over the Plateau is the greatest in China. Tang (1979) analyzed the flow pattern at low and high levels over the Plateau in July, and found that the weather systems also have intense diurnal variation. Figure 6 illustrates the difference in mean geopotential heights between 1800 and 0600 (Lhasa local time) in July at (a) 600 hPa, and (b) 100 hPa. Both pressure systems at low and high

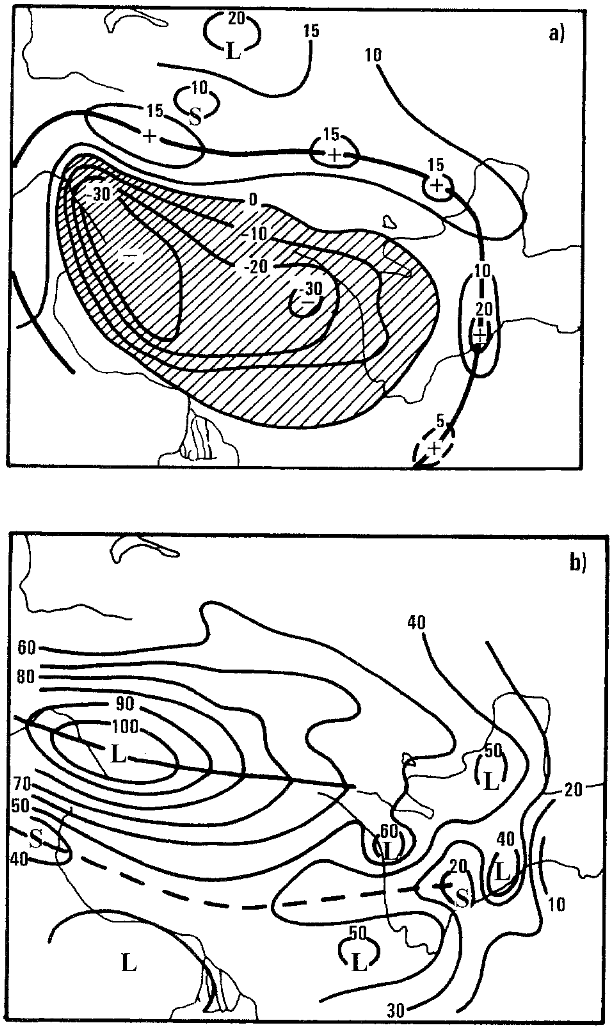

Fig. 6. Monthly mean geopotential height difference between 1800 and 0600 (Lhasa local time) in July. (a) 600 hPa; (b) 100 hPa. Adopted from Tang (1979)

levels are stronger in the afternoon. This is consistent with the strong diurnal variation in convective activity (Fig. 5c). An interesting pronounced feature seen in Fig. 6 is that there are two maximum centers of diurnal variation situated over the western and central-eastern Plateau, respectively. This noticeable phenomenon is also consistent with the existence of the two ascending motion (Fig. 3) and humidity centers over the eastern and western Plateau pointed out by Yang et al. (1992a). From the above discussion, it becomes obvious that small-scale convection is very important for the maintenance of large-scale circulation over the Tibetan Plateau in summer.

5. The Influence of the Elevated Heat Source over the Plateau on the Atmospheric Circulation in Summer

The fact that the heat source has a large effect on the atmospheric circulation has long been known (e.g., Smagorinsky, 1953). This should be especially true for the heat source over the Plateau mainly because of its size, height and latitudinal position. Firstly, the Plateau is situated in the subtropical belt where the prevailing current is weak in summer, thus the heat generated there will not be carried away quickly. Secondary, the heat produced there is directly added to the middle troposphere and used only to heat half of the atmosphere. Further, in summer, as discussed in the last section, the very strong convergence near the surface and divergence in the upper troposphere act as important vorticity source near the surface and sink in the upper troposphere. All these facts must have large impacts on general circulation and climate. In summer, the persistent existence of a surface low with strong convergence and an upper level high with strong divergence over the Plateau becomes an unique weather pattern in the world. This is an example of the impacts on regional scale. Example of the impacts on global scale will be discussed in the following.

As discussed in previous sections, in summer over the Plateau the ensemble of enormous convective cells plays an important role in balancing the strong positive vorticity generation in lower layers and negative vorticity generation in upper layers. In any single convective cell, the upward current can be compensated for by the downward motion surrounding it. But for the whole Plateau, there exists a strong mean upward current resulting from these numerous cells, with two centers, respectively, over its western and eastern parts as shown in Fig. 3. Where will the large amount of rising air go? And how will this influence the global circulation? To answer this question, we used the NCAR/NCEP monthly mean data ranging from 1980 to 1995 and plotted the longitudinal and meridional circulation cross-sections which pass across the Plateau. From the meridional circulation along 90° E in July (Fig. 7a) we see that lower layer flows converge from both north and south towards the Plateau to form strongly ascending air. Such ascending air flows

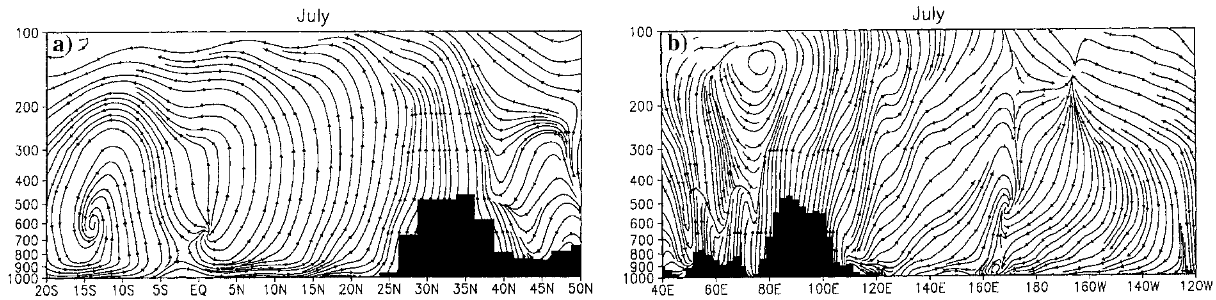

Fig. 7. The vertical cross-sections of streamline along 90° E (a) and 30° N (b) in July averaged from 1980–1995. The shaded area presents orography

southward from the Plateau in the upper troposphere, then goes across the equator, merges with the upper branch of the southern hemisphere Hadley Cell, and descends in the Southern subtropics, forming the well-known Hadley-monsoonal meridional circulation. In fact, this monsoonal meridional circulation exists in the area from 55° E to 140° E, rather than being confined to the longitudes of the Plateau. An interesting phenomenon is that the compensating descent to the north of the Plateau is observed between 40 and 47° N, where the Takla Makan and Dzungaria Desert are located. From May to June this descending air and the ascending air over the Plateau from a closed smaller cell just over the northern flank of the Plateau (figures not shown). The results obtained here agree well with the previous findings of Ye and Yang (1979).

Figure 7b is the vertical cross-section of the mean July longitudinal circulation along 30° N. The ascending air coming from the surface of the Plateau flows eastward, then descends east of 170° E, constituting an intense east-west circulation as noted by Krishnamurti (1971). The ascending air over the Plateau also flows westward, then descends over Afghanistan, Iran, and Saudi Arabia, leading to the formation of a dry climate there. Similar results have also been obtained by Yang et al. (1992b) for the year 1979. A recent study on the dynamics of deserts of Rodwell and Hoskins (1996) shows that the latent heat release due to condensation of the Asian monsoon which is associated with the Tibetan Plateau forcing can excite a Rossby wave pattern with its rear descending motion just over the desert areas in Middle East Asia and North Africa.

Numerical experiments of Hahn and Manabe (1975) and of Zhu (refer to Wu et al., 1996) show that without the presence of the Tibetan Plateau in a GCM, there should be no occurrence of the Asian monsoon, and the main rain band over Asia should shift southward. Recently Wu et al. (1997a) conducted another numerical experiment using the Global Ocean-Atmosphere-Land System climate model of the State Key Lab of Atmospheric Sciences and Geophysical Fluid Dynamics (GOALS/LASG) (Wu et al., 1997b). In this experiment the Tibetan Plateau was kept unchanged but the surface sensible heating was removed from the atmospheric thermodynamic equation. Similar results as in Hahn and Manabe's experiment, in which orography was removed, were obtained. This suggests that not only the mechanical forcing but also the elevated surface heating of the Plateau is important for the occurrence of the Asian monsoon. To further illustrate the influence of sensible heating of the Plateau, Wu et al. (1997a) also calculated the differences of the 200 hPa stream function between experiments without and with surface sensible heating on the Tibetan Plateau. The results for the July mean are shown in Fig. 8. In the control experiment (CON) in which the SH on the Plateau is kept, the simulated 200 hPa stream field (Fig. 8a) is close to the observed July climate. When the surface sensible heating on the Plateau is removed, the difference field presented in Fig. 8b shows that a strong cyclonic center appears just over the Plateau, and a Rossby wave ray with zonal wave number 2 is generated over the globe. The wave ray emanating from the Plateau propagates eastward. It bends southward over the Northern Pacific, enters the tropics through the westerlies at the bottom of the

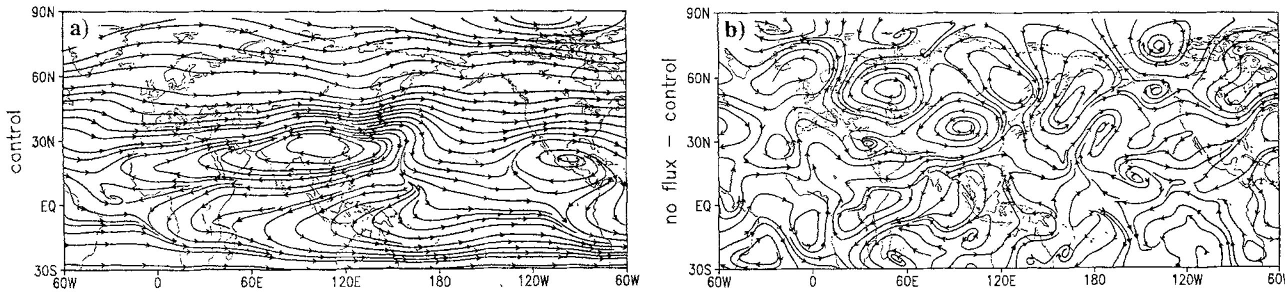

Fig. 8. The 10-year mean July stream field at 200 hPa in the control run (CON) simulated by using the GOALS/LASG climate model (a), and its difference between the experiments without and with surface sensible heating over the Tibetan Plateau region where the absolute elevation is above 3 km

tropical upper-tropospheric trough (TUTT) there (see Fig. 8a), then propagates into the southern hemisphere. This wave ray splits over the Northern Pacific. The northern branch propagates further eastward. It passes over Alaska and Baffin Bay, then bends southward over the North Atlantic, and disappears over the north of Cap Verde due to the obstruction of the axis of the subtropical high (where u = 0) and the easterlies to its south. Wherever this wave reaches, the distribution of troughs and ridges of the stream field is disturbed (figures not shown). The above results therefore suggest further that the influence of the anomalous elevated surface heating of the Tibetan Plateau on climate can be global.

6. The Influence of the Tibetan Plateau on Seasonal Variation

In the 50's, Ye et al. (1959) found that the seasonal transition from winter to summer of the atmospheric circulation over the South Asia is abrupt: within a few days the mid-latitude westerly jet, the axis of the subtropical anticyclone, and the tropical easterlies all shift northward abruptly by 5 to 10 degrees of latitude. Zeng et al. (1988) have simulated such kind of abrupt seasonal transition by using the IAP GCM. To understand the cause of such abrupt seasonal change, Wu et al. (1997a) designed a set of numerical experiment. They replaced the ocean component of the GOALS/LASG climate model by the prescribed observed sea surface temperature (SST) ranging from Jan. 1, 1979 to Dec. 31, 1988, and integrated the model for the ten-year period to conduct a control run (CON). Then they removed the surface sensible heating to the atmosphere in the model, kept other parameters unchanged, and

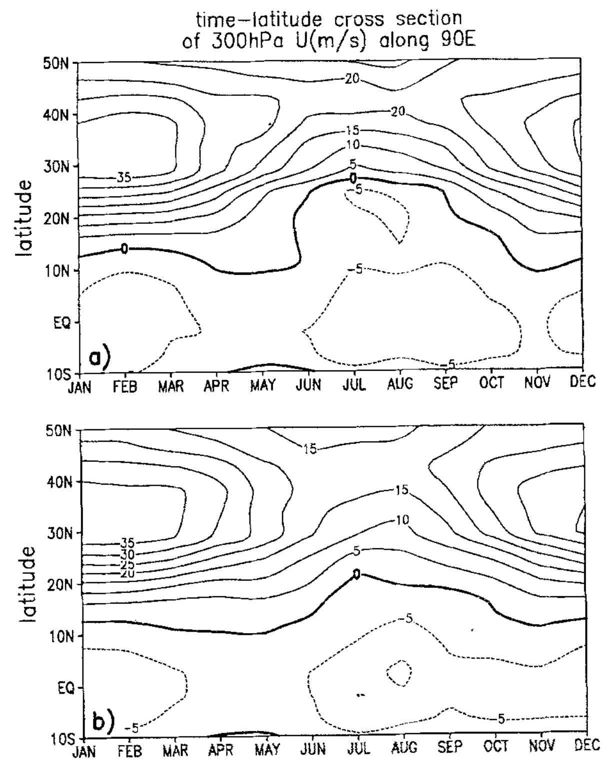

Fig. 9. The 10-year mean seasonal variation of the zonal wind component along 90° E in the CON run (a), and in a parallel experiment (NSH) in which the surface sensible heating is removed from the plateau region where the absolute elevation is above 3 km (b). Unit is in ms^{-1}

integrated the model for the same period to conduct a no-sensible-heating run (NSH). Figure 9 shows the ten-year mean seasonal variations of the westerly wind component at 300 hPa along 90° E obtained from the two runs, respectively. In the CON run (Fig. 9a) in winter and early spring, easterlies are maintained near the equator, a westerly jet, at 35° N, and the axis of the subtropical anticyclone ($u = 0$) at 10° N. At the end of May and the beginning of June, the axis of the subtropical anticyclone suddenly shifts northward from 10 to 27° N, and the westerly jet retreats correspondingly from 35 to 45° N. These results agree fairly well with those obtained by Ye et al. (1959), and indicate that the model is capable of simulating the mean climate state. On the other hand in the NSH run (Fig. 9b), the axis of the subtropical anticyclone evolves from 10° N in May to 20° N in July over a period of about two months and varying smoothly. This then suggests that the abrupt seasonal change of such Asian circulation system may be attributed mainly to the elevated surface heating of the Tibetan Plateau.

Following Li and Yanai (1996), Wu et al. (1997a) adopted the difference between 30° and 5° N of the mean temperature averaged from 200 to 500 hPa (ΔT) to present the thermal characteristics of the upper troposphere along the tropical and subtropical zones. A positive ΔT can be used to indicate the Asian monsoon period. By using the NCAR/NCEP monthly mean reanalysis data from 1980 to 1995, the monsoon period is estimated to persist from the middle of May to September (Fig. 10a). By using the model output of the CON run mentioned above, the simulated monsoon period is from the middle of May to the middle September (Fig. 10b), fairly close to the analysis based on the reanalysis data shown in Fig. 10a. In the NSH run, however, the monsoon starts not till early June, and ends as early as in the middle of August (Fig. 10c). The monsoon period is shortened by half in the NSH run compared to that in the CON run and in the "observations". In addition, the most persistent longitude of the monsoon period also shifts from near the Plateau region in the CON run to about 120° E in the NSH run. These results imply that the elevated surface heating of the Tibetan Plateau is very important for the location and persistence of the Asian monsoon.

It has long been recognized that the Tibetan Plateau forcing is also important for the Asian monsoon onset. Yanai et al. (1992) conducted a systematic review on this subject, and examined the warming process of the upper troposphere in 1979. They reported that the temperature increase over the eastern Plateau during the Southern China Sea (SCS) monsoon onset in May was mainly the result of diabatic heating,

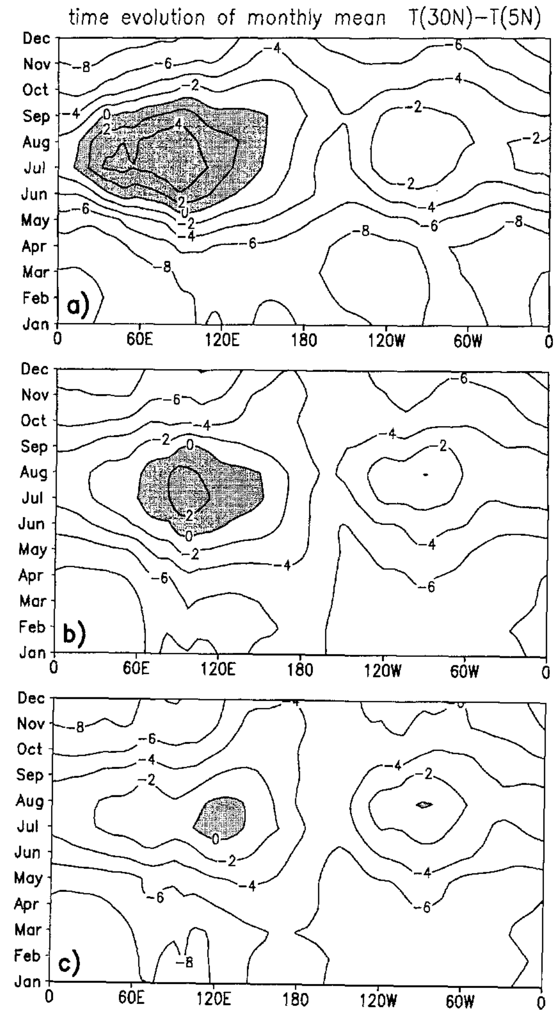

Fig. 10. The longitudinal distribution of the seasonal variation of the difference between 30 and 5° N in the mean temperature averaged from 500 to 200 hPa. Shaded region indicates positive difference and represents the summer monsoon period. Unit is in °C (a) 1980–1995 mean observation, produced from the NCAR/NCEP reanalysis data; (b) 10-year mean in the CON experiment; and (c) 10-year mean in the NSH experiment

whereas that over the Iran-Afghanistan-western Plateau region leading to the Indian monsoon onset in June was caused by intense subsidence. Chang and Chen (1995) made a brief review of possible mechanisms linking the Asian monsoon onset to the Plateau forcing. They also suggested that due to the blocking effect of the Plateau, the early monsoon onset occurs over the SCS region because only over this region can the mid-latitude trough meet with tropical disturbances in springtime. Recently Wu and Zhang (1998) employed the ECMWF/TOGA complementary data set, and diagnosed the evolution of Asian circulation during the seasonal transition from winter to summer in 1989. They found that both the mechanical forcing and thermal forcing of the Tibetan plateau are important for the Asian monsoon onset, and the whole onset period can be divided into three stages: the Bay of Bengal (BOB) monsoon onset in early May, the SCS monsoon onset in the middle of May, and the Indian monsoon onset in early June. Mechanically, the low-layer westerlies impinge upon the Plateau at its southwest corner in early spring and go around its southern periphery, forming a low-layer anticyclonic circulation over India and the Arabian Sea, and a cyclonic circulation, the so-called Burma trough, over the northern part of the Bay of Bengal. This then provides the background for the early onset of the Asian monsoon to occur over the eastern coastal region of the Bay of Bengal, and is in agreement with the early findings of Yin (1949) that the earlier rainy season in the South Asia region starts in Burma. Thermally, the remarkable elevated surface sensible heating of the atmosphere over the Plateau started in March 1989. Afterwards the air aloft is heated stably at a rate of 2 to 4 °C per day by SH. Due to advection, however, the air column over the eastern part of the Plateau is warmer than that over its western part by several degrees. This is also the case in 1979 as reported by Yanai et al. (1992, Fig. 20a). The warmer air over the eastern part of Tibet causes the southerly inflow from the Bay of Bengal towards the Plateau to develop in the lower troposphere, which brings plenty of water vapor into the northern part of the Bay of Bengal and the surrounding inland area, a kind of circulation in favour of the early monsoon onset there. The development of a surface cyclone and upper layer anticyclone over Burma during BOB monsoon onset then causes the development of southerlies and convergence in the lower troposphere, and northerlies and divergence in the upper troposphere over the SCS area. The SCS monsoon onset then follows. Afterwards, only when the whole flow pattern has shifted westward, can the Indian monsoon onset occur, and that is already in early June. The three-stage feature of the Asian monsoon onset can be seen clearly from the Hovmoeller diagram of the longitudinal distribution of pentad mean outgoing longwave radiation (OLR) and wind vector at 850 hPa averaged over the latitude range from 10° to 20° N as shown in Fig. 11. The shaded area in the figure denotes where the OLR is less than 220 W m^{-2} which indicates the region of development of deep convection. It looks like a cactus tree with the top three branches accompanied by the development of southerly or southwesterly winds. These three branches appear respectively in early May in the BOB region, on May 20 in the SCS region, and on June 10 in the India region, presenting the three-stage feature of the Asian monsoon onset in 1989. By examining recent historical records of 10 years, it is found that this feature exists every year (Zhang and Wu, to be published in the Chinese Journal of the Atmospheric Sciences). It is evident that the onset of the Asian monsoon is related directly to the mechanical as well as thermal forcing of the Tibetan Plateau.

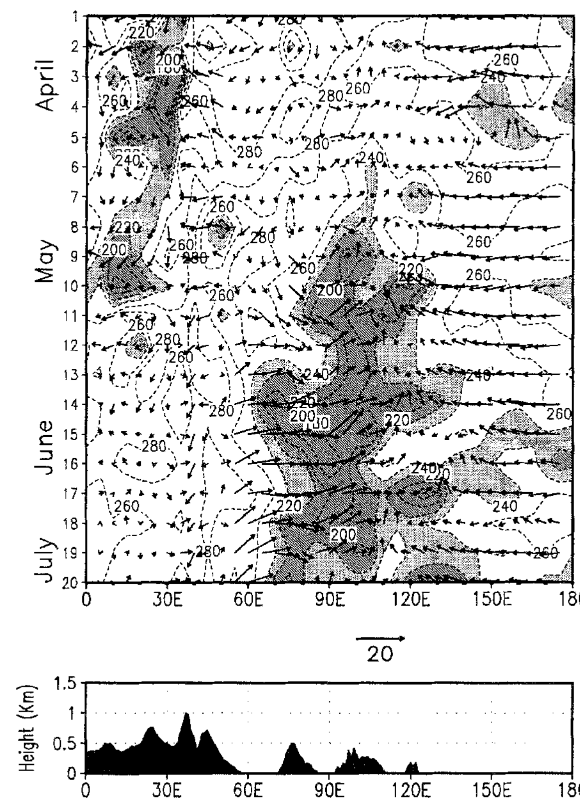

Fig. 11. The evolutions in the seasonal transition period of 1989 of the longitudinal distribution of 850 hPa wind and outgoing long-wave radiation (OLR) meridionally averaged from 10 to 20° N. Contour interval is 20 W m^{-2}

7. Minimum Ozone Center over the Tibetan Plateau

Using the TOMS data from 1979 to 1991 provided by the Goddard Space Flight Center/NASA, Zhou et al. (1995) analysed the distribution of total ozone concentration over China. The results shown in Fig. 12a indicate a nearly zonal distribution in January. From June to September, there appears a minimum center of total ozone concentration just over the Tibetan Plateau (Fig. 12b). The intensity of this minimum center decreases from October onwards. They also calculated the monthly mean total ozone concentration averaged over the Plateau region, and compared it with that over the eastern China within the same longitude domain, the relative difference is presented in Fig. 13. Although all the year round the total concentration over the Plateau is lower than that over eastern China, the difference in winter and spring is less than 4%, while in the summer months the difference is over 10%, with a maximum of 11% appearing in June. Although the intensity of this minimum ozone center is not as strong as the "ozone hole" over the Antarctica region, it is an enormous phenomenon in the northern hemisphere. Zhou and his colleagues linked the appearance of this minimum center in summer to the strong convective activity developed over the Tibetan Plateau. They postulated that as such activity brings air with low concentration of ozone from the lower troposphere to the upper troposphere and stratosphere, it also carries upward chemical pollutants which are collected from the surrounding region by the inflow in the lower troposphere. These chemical pollutants may also deplete the ozone concentration in high layers of the atmosphere due to some physical or chemical processes, resulting in low ozone concentrations there as well. In any case, the formation of the minimum ozone center over the Plateau in summer must be related to the particular convective activities in the area. Recently Zhou et al. (1997 personal communication) performed a numerical experiment proving this idea.

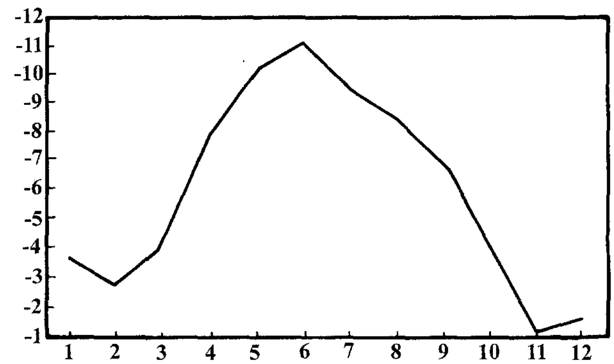

Fig. 13. The annual variation of the relative difference in total ozone concentration between the area over the Tibetan Plateau and that over the eastern China with same latitude rages. Adopted from Zhou et al. (1995)

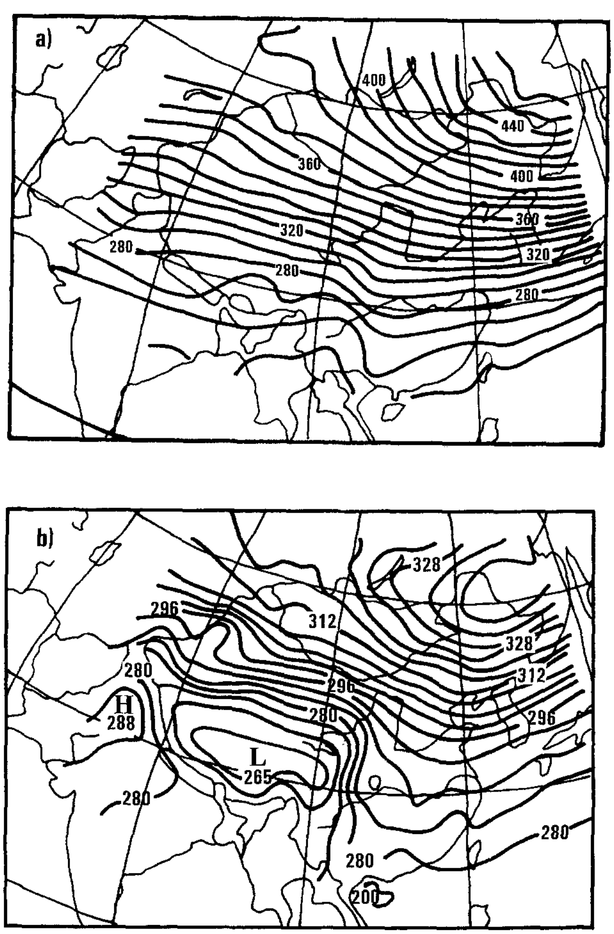

Fig. 12. The distributions of total ozone concentration in January (a) and September (b) averaged from 1979 to 1991. Adopted from Zhou et al. (1995). Unit is in Dobson

8. Summary and Discussion

Due to its particular elevation, vast area, and clear air, the Tibetan Plateau receives a large amount of solar radiation, leading to a very unstable stratification in the lower layers of the atmosphere in summer. Up to 2 km from the surface, the temperature lapse rate can usually be super-dry-adiabatic, resulting in the strong development of convection. This intense convective activity leads to a deep mixed layer with equivalent potential temperature (θ_e) reaching 400 hPa (Luo and Yanai, 1984; Song et al., 1984). The mean intensity of convergence at low levels and that of divergence at high levels can reach $10^{-6}\,s^{-1}$ over the Plateau (Ye et al., 1957; Ye and Yang, 1979; Flohn, 1968). Therefore, a mean ascending current with a magnitude of cm/s must exist there. Two maximum centers upward motion are located in the eastern and western Plateau, respectively (Yang et al., 1992b).

This intense convective activity is essential for maintaining prevailing low-pressure systems at low levels and high-pressure systems at high levels in summer. Ye et al. (1974) and Yang et al. (1992c) found that the production of vorticity can not be balanced by the mechanisms of advection of the large-scale circulation alone, and the role of strong convective activity over the Plateau is very important in the vorticity balance. This feature is different from that in the tropical and middle latitude areas. As discussed above, the maintenance of the pressure systems over the Plateau in summer is in a large extent attributed to its particular thermodynamic structure and the consequent strong convective activity.

A mean large-scale upward motion persists almost all the day over the Plateau due to stable large-scale convergence at lower levels and divergence in the upper troposphere and lower stratosphere (Yang et al., 1992c). This is another important feature of the atmospheric circulation over the Tibetan Plateau in summer. The large amount of air from the upward motion diverges in all directions at high levels. One branch flows eastward to the eastern Pacific Ocean and descends there, forming a large-scale east-west vertical circulation north of 30° N. Another branch current flows westward to Iran and even Africa and subsides there. This is related to the formation of the dry climate in these areas. Part of the current flows southward across the equator and descends in the Southern Hemisphere. This, together with the strong monsoon circulation, has an important influence on the climate of the Southern Hemisphere (Ye and Yang, 1979). Furthermore, the anomalous surface heating can affect the global circulation pattern via energy dispersion on spherical surface. Thus, the heating of the Plateau affects not only the weather in monsoon regions and in the subtropics, but also affects the global general circulation. Furthermore, strong surface sensible heat fluxes occur over the Tibetan Plateau from early spring onwards, which can warm up the air aloft persistently by 2 to 4 °C per day. With the westerly advection before monsoon onset, the heat source over the Plateau makes the air column over its eastern part of the Plateau warmer than that over its western part. A southerly inflow towards the Plateau thus develops over the eastern part of the Bay of Bengal. This, together with the mechanically forced Burma trough, causes the onset of the Asia monsoon earlier along the eastern coast of the Bay of Bengal, which is then followed by monsoon onsets over the South China sea and India.

The elevated heating of the Plateau has also important impacts on the seasonal variation of the atmospheric circulation. In a numerical experiment in which surface sensible heating from the Tibetan Plateau is removed, the northward march from winter to summer of the easterlies, of the axis of the subtropical high, and of the mid-latitude westerly jet in the Tibetan Plateau area become very slow and smooth. Besides, the Asian monsoon period is shortened greatly. These results suggest that the elevated heating of the Plateau is essential for the occurrence of the abrupt seasonal change of the general circulation over Asia, and for the persistent maintenance of the Asian summer monsoon.

The strong development in summer of low layer convergence and upward motion over the Plateau transport from the surface to the upper atmosphere not only the air with low concentration of ozone but also chemical pollutants. These will deplete the ozone layer over the plateau through chemical and physical processes, thus producing a region with relative low ozone concentration over the Tibetan Plateau in summer.

Although great achievements have been obtained through the combined efforts of atmospheric scientists over the world, due to the scarce observation network and the complicated terrain and topography, many features of the impacts of the Plateau, particularly those associated with land surface processes, including soil type, soil moisture, vegetation type, etc., and their impacts on climate are little known. More accurate theories to explain the particular physical exchange processes at the surface over such a highland are needed. Further, more accurate estimates of the intensity of the heat source are also needed. To explore these, a new national research program on the influence on regional and global climate of the physical processes at the surface of the Tibetan Plateau has been organized in China. A field meteorological observation experiment in Tibet (TIPEX) is planned for the period May to August, 1998. We are looking forward to gaining new insights into this issue in the near future.

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